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# **Estimation of soil redistribution rates due to snow cover related processes in a mountainous area (Valle d'Aosta, NW Italy)**

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## **1 Introduction**

Soils in mountain areas are fragile and often scarcely developed, mainly because of slope steepness and extreme climate conditions, and they are widely affected by erosion processes and soil slip involving mainly superficial horizons. Main drivers for soil erosion are topography, land cover, soil texture, rainfall patterns and land use (Wischmeier and Smith, 1978). Soil erosion is mainly linked to rainfall runoff processes, but, in mountain areas, also snowmelt runoff and snow movement related processes may be important factors that can enhance soil erosion (Konz et al., 2009). The snow movements which could involve the soil are: a) the snow gliding (slow glide movement of the snow cover on smooth or wet ground, that can lead to the formation of glide cracks), and b) the ground avalanches (in particular the full-depth slab avalanches or so called gliding avalanches, occurring when the snow gliding changes to a rapid movement. Also the snowmelt period can contribute significantly to the total annual runoff and sediment yield, constituted mainly by fine particles (< 2 mm) (Lana-Renault et al., 2011). However, the role and importance of snow as a soil erosion agent are not well understood yet. In fact, the snow movements can exert considerable erosive forces on soils; in particular avalanches can transport consistent amounts of debris, especially when they involve the whole depth of snow or run onto snow-free areas (Luckman, 1978; Freppaz et al., 2006; Freppaz et al., 2010; Confortola et al., 2011). The rock and soil material, transported by snow movements, origins from the erosion of the underlying soil and bedrock in the release zone and along the track area (Gardner, 1983; Jomelli and Bertran, 2001). Deposits of debris from snow avalanches are common in mountain environments indicating the importance of avalanches as specific geomorphic agents. Mapping and quantification of soil erosion under different land-use conditions have been evaluated in many projects for agricultural soils in lowlands. However, only few studies exist on soil erosion measurement and quantification in low-

mountain ranges (e.g. Nearing et al., 1999; Leser et al., 2002; Gabriels et al., 2003; Prasuhn et al., 2007) or in alpine environments characterized by specific climatic and topographic conditions (Felix and Johannes, 1995; Descroix and Mathys, 2003; Isselin-Nondedeu and Bedecarrats, 2007; Konz et al., 2009, 2010, 2011; Meusburger et al., 2010). Moreover, most of the studies focus on soil-erosion measurement during the vegetation period, while only few works deal with soil erosion caused by snow movements related processes (Ackroyd, 1987; Bell et al., 1990; Kohl et al., 2001; Heckmann et al., 2005; Konz et al., 2009, 2011; Freppaz et al., 2010). Hence, more data on soil erosion in alpine regions, for longer time periods and including all the seasons, are needed (Konz et al., 2009). Taking into account the soil redistribution process related to the “snowy season”, conventionally, it can be assessed by the measurement of sediments transported in the runout zones by snow movements (snow gliding and ground avalanches), considering both the debris and the fine material (e.g. Ackroyd, 1987; Bell et al., 1990; Kohl et al., 2001; Heckmann et al., 2005; Freppaz et al., 2010). On the other side the yearly soil erosion in mountain areas can be measured by isotopic techniques based on the use of fallout radionuclides “FRNs” such as  $^{137}\text{Cs}$ ,  $^7\text{Be}$  and  $^{210}\text{Pb}$  (Zapata, 2002; Mabit et al., 2008a). “FRNs”, in particular  $^{137}\text{Cs}$ , have proven to be very powerful tracers of soil movements within the landscape, and this methodology can complement conventional approaches (Mabit and Fulajtar, 2007). The presence of  $^{137}\text{Cs}$  in the soil is either due to nuclear weapon testing (‘60s) and/or to the Chernobyl reactor accident in 1986. Input of  $^{137}\text{Cs}$  through the Chernobyl reactor accident was highly dependent on the rainfall pattern which caused high (kilometre to regional scale) heterogeneity in  $^{137}\text{Cs}$  distribution (Higgitt et al., 1992; Renaud et al., 2003). For a small catchment or single hill slopes a homogeneous rainfall pattern can be assumed. After deposition,  $^{137}\text{Cs}$  is rapidly and tightly bound to fine soil particles and its redistribution is mainly caused by soil erosion as it moves with soil particles (e.g. Bonnett, 1990; Ritchie and McHenry, 1990). The use of  $^{137}\text{Cs}$  measurements to estimate rates of erosion and deposition is based on comparison of the inventories at individual sampling points with a reference inventory, representing the local fallout input. The reference site is expected to show neither erosion nor deposition. A measured inventory for an individual sampling point less than the reference value is indicative of erosion, whereas an inventory greater than reference value is indicative of deposition (Walling and He, 1999). To convert  $^{137}\text{Cs}$  measurements to quantitative estimates of erosion and deposition rates, many different methods exist, including both empirical relationships, and theoretical models and accounting procedures (Walling and He, 1999). The aim of this study is to quantify soil erosion and deposition rates due to snow movements related processes (snow gliding and ground avalanches) by conventional field measurements in a snow deposition area. Moreover the yearly soil redistribution rates in different sites of the study area will be assessed with the Cs-

137 method, which yields the cumulative net soil loss/gain since 1986, including all the erosion processes. Our hypothesis is that the combination of the two approaches will provide information on the relative importance of soil erosion related to snow movements in comparison to runoff processes.

## **2 Materials and methods**

### **2.1 Study area**

The watershed investigated (24.6 ha) is located in the north-western part of the Valle d'Aosta Region (NW-Italy), very close to the Mont Blanc Massif (4810 m a.s.l.) in the municipality of Courmayeur. The study area is an avalanche site called "Torrent des Marais", running on a west exposed slope, from 2115 m to 1250 m a.s.l. (Fig. 1). In this site, besides the dry snow avalanches, also wet snow, full-depth slab avalanches (gliding avalanches) are typical during the spring time, and sometimes also in late autumn; in fact, the release and track areas are often characterized by intense snow gliding and glide cracks formation. All these processes are related with soil erosion. Full-depth slab avalanches in particular often involve the soil surface with the removal, subsequent transport and accumulation, in the deposition areas, of the upper horizons mixed with rocks and plant material. According to the Avalanche Cadastre of the Valle d'Aosta Region, in the last four years the frequency of gliding avalanches in the investigated area seemed to have increased: for each of the last four winters we had one documented event, while previous data recorded in the Regional Avalanche Cadastre, within the last four decades, indicate only few ground, wet snow avalanches. The last events took place during winter seasons with differing snow conditions (2007/2008: low snowfall; 2008/2009 and 2009/2010: high snowfall; 2010/2011: low snowfall) and the last one occurred in autumn instead of spring. However we have to consider that, historically, the Avalanche Cadastre of the Region Valle d'Aosta was used to record mainly the events involving human activities; just in the last six years the avalanche cadastre have been improved and aimed to collect all the avalanche events occurring each season. Moreover specific investigations carried out in the area, based on the folk memory, let us assume that the frequency of this kind of avalanche events (full-depth slabs of wet snow) in the last decades was constant and almost yearly.

Considering a larger catchment scale, the described study area, according to the Avalanche Cadastre, is in the middle between two big districts: Val Ferret and Morgex . From the overlap (ArcGis 9 - ArcMap Version 9.2) between/of avalanche areas, registered in the Avalanche Cadastre from '70s, and the vegetation cover groups, got from the land use map available for the Valle d'Aosta Region, we obtained that 34% of the avalanche release and track areas is covered by alpine meadows, the most prone to ground avalanches and related soil erosion processes because of its

lowest roughness, 2% by hayfields and 8% by dwarf shrubs (in order of increasing roughness coefficient) (Leitinger et al., 2006). The avalanche release area (RA) is characterized by mean steepness of 30° and is covered by abandoned pastures. The track area (TA) becomes much channelled (>35° of inclination) and its coverage is composed mainly by grass. In the steepest areas, rocks and bare soil are frequent. However, the main deposition area (LDA) has a decreasing steepness and it ends on an avalanche shed, constructed in the 70's to protect the regional road. Besides this main deposition area, there's a secondary smaller deposition area which is located at upper elevation (UDA) at 2015 m a.s.l., just below the release zone. A portion of the release area is covered with snow bridges (active avalanche defence structures) since 1974, and in this area (SB) the vegetation cover is mainly characterized by dwarf shrubs (*Vaccinium myrtillus* and *Rhododendron ferrugineum*) and several larch (*Larix decidua*) seedlings. Just above the snow bridge, at 2110 m a.s.l., there's a flat area (ASB), where the inclination gently decreases, and then the slope above is again steep up to the top. Meteorological data are continuously recorded by an automatic weather station (2076 m a.s.l.) of the Ufficio Centro Funzionale (UCF) – Valle d'Aosta Region (VDA), located very close to the study area. The long-term yearly mean precipitation recorded at this station is 840 mm yr<sup>-1</sup> (period 1995–2010) and the mean annual air temperature is +2.8 °C (period 1993–2010) (Source: UCF-VDA). The average cumulative annual snowfall is 275 cm at 1250 m a.s.l. (period 1937–1995) and about 450 cm at 2000 m a.s.l. (Source: SMS, 2003). The bedrock is constituted mainly by black argillic schists and calcareous sandstones and, in some places, by porphyritic granites. During the first decade of May, 1986 (Chernobyl accident), north-western Italy was disturbed by both wet and dry radioactive fallout and the study area was affected by a precipitation estimated in 5–10 mm of rain. After a sampling campaign conducted in the Valle d'Aosta Region in 2001–2004 by ARPA Valle d'Aosta (Regional Environmental Agency), the <sup>137</sup>Cs concentration in the soil was estimated to be in the range 0–6000 Bq m<sup>-2</sup> (Agnesod et al., 2006). Moreover in the close Piemonte Region, it was estimated that the contribution of Chernobyl wet deposition was the major part of the global inventory (84 %), while 9% was attributable to pre-Chernobyl caesium (derived from atomic bomb testing in the atmosphere) and 7% to Chernobyl dry deposition (Facchinelli et al., 2002). Because of the proximity between the two Regions, we can consider these data most likely attributable also to our study area. This is an important assumption because if the Chernobyl input is considerably greater than the input of <sup>137</sup>Cs associated with “60's bomb fallout”, then the <sup>137</sup>Cs measurements will primarily reflect erosion occurring after 1986.

## 2.2 Methodology

### 2.2.1 Snow sampling in the avalanche deposition area

Two full-depth slab avalanche events were considered in this work: (a) one during winter season 2008–2009 (1 March 2009) and (b) the other in winter 2009–2010 (17 March 2010). Both snow avalanches caused soil erosion, with the formation of very dirty snow accumulations in the deposition zones (LDA-UDA) (Fig. 2). The release and track areas were calculated with GIS, on the basis of the georeferenced pictures taken after the avalanche events. The deposit area, by GPS measurement of the perimeter, and the snow depth, for calculating the volume, were determined in the field as soon as the safety of the area was judged adequate. Snow depths ( $n=6$ ) were measured by a 3.5 m snow probe. For the purpose of sediment estimation, each avalanche deposit was surveyed two times: 17 March and 19 May during winter 2009; 2 April and 17 May during winter 2010. Snow in the main avalanche deposition zone (LDA) was sampled according to a gridded design. During winter 2009 the sampling points were distributed on a  $7\times 7$  m grid square during the first sampling campaign (17 March ( $n = 58$ )) then on a  $15\times 15$  m grid square during the second campaign (19 May ( $n = 10$ )). Sampling in 2010 was conducted at a lower spatial resolution, using a  $20\times 20$  m grid square in both sampling events (2 April ( $n = 8$ ), 17 May ( $n = 6$ )). The sampling points were mapped by GPS (accuracy  $<5$  m). We chose to sample the surface and the inner body of the avalanche deposit in order to investigate the vertical distribution of sediments. Superficial samples were collected on a constant area, by a wooden made rectangular mask of  $0.15\times 0.15$  m in order to: a) make easier and faster the sampling campaign as the avalanche run out area was an unsafe zone; b) try to avoid the extreme “topographical” irregularity of the deposit surface. About the sampling depth a constant of 0.02 m was estimated. In the inner body snow samples were collected at 0.20 m below the surface using a plastic core (fixed volume of  $0.15\text{ dm}^3$ ). Moreover a snow pit was dug down to the ground in the avalanche deposit during spring 2009 and 5 samples ( $0.15\text{ dm}^3$ ) were collected in order to analyze the debris distribution along the deposit profile (Fig. 2). During the sampling we observed that most part of the sediment was represented by “soil”, a mixture of small mineral and organic particles. We excluded large organic debris and rocks with a diameter greater than 5 cm.

### **2.2.2 Estimates of snow related soil redistribution rates by avalanche deposit data**

The samples were melted and filtered through a  $0.45\text{ }\mu\text{m}$  filter, using a vacuum pump. The filtered material was dried ( $40\text{ }^{\circ}\text{C}$ ) and weighted to obtain the concentration of sediment ( $C$ ) that was used to estimate the sediment load in the run-out area by multiplying with the avalanche deposit volume ( $V$ ). The sediment load was calculated for both the deposit surface layer ( $V = \text{deposit surface area} \times \text{snow depth of the superficial layer (0.02 m)}$ ) and for the rest of the snow deposit ( $V = \text{deposit surface area} \times \text{deposit mean depth}$ ). The average soil deposition rate was calculated dividing the

total sediment load of each event by the deposition area. This mass accumulation rate was converted into a rate of soil accretion dividing the soil deposition rate by the unconsolidated debris density of  $1200 \text{ kg m}^{-3}$  (Freppaz et al., 2010). The average soil erosion rate for each event was calculated dividing the total sediment load by the sum of avalanche release and track areas.

### **2.2.3 Soil sampling along the avalanche path**

Soil samples were collected during summer season 2010, using a 72 mm diameter soil core sampler, (Giddings Machine Company, Windsor, CO, USA). The site for the reference inventory (RS), that represents the original  $^{137}\text{Cs}$  activity without soil redistribution processes, was selected very close to the study area in a flat and undisturbed position at 2000 m a.s.l. We took 11 soil profiles from the reference site: three soil samples, which were a bulk of three replicates taken within  $1 \text{ m}^2$ , were used to determine average profile distribution and maximum depth of  $^{137}\text{Cs}$ , while eight samples were used to investigate its spatial variability and to determine the caesium-137 base line. For estimating redistribution rates soil cores were collected along altitudinal transects in the three sites SB, RA and TA; just for the track area (TA) we chose to sample two transects (TA1 and TA2) instead of one because of the wider extension of this area compared to the others. For each transect 5 cores were taken every 15 m of distance. From these cores the upper 9 cm were separated from the lower soil profile, in order to differentiate the physico-chemical properties of topsoil from the subsoil. In the lower main avalanche deposition area (LDA) 7 samples were collected and bulked together every 5 cm, while three sites were sampled in the upper deposition area (UDA) and other three in the flat area above the snow bridge (ASB) (Fig. 1).

### **2.2.4 Soil samples pre-treatment and laboratory analyses**

In total 62 soil cores were sampled, and from them 136 sub-samples were oven-dried at  $40^\circ\text{C}$ , lightly ground and sieved ( $<2 \text{ mm}$ ). An aliquot of each sample was put into a 25-mL container (6.5 cm diameter; Semadeni25) and analyzed for 8 h with a Li-drifted Ge detector (GeLi; PrincetonGamma-Tech, Princeton, NJ, USA) at the Department for Physics and Astronomy, University of Basel. The size of the detector was 48 mm in diameter and 50 mm in length. The relative efficiency was 18.7 %. In order to reduce the amount of radiation from background sources into the environment, the samples were shielded by 4-cm-thick lead during measurement. The  $^{137}\text{Cs}$  activity concentrations were determined using the Inter Winner 5 gamma spectroscopy software (Ortec, Oak Ridge, TN, USA). The energy calibration of the GeLi detector was done using a Eu-152 multi-source with peak line positions at 117.6, 347.6, 773.5, 1108.0 and 1408.9 keV. For efficiency calibration, three reference-samples provided by H. Surbeck (University of Neuchâtel)

enriched with known activities of U-238, Th-232 and K-40, were used. These calibration samples were of the same geometry and a comparable density as our analysed soil samples. The resulting measurement uncertainty on  $^{137}\text{Cs}$  peak area (at 661 keV) was lower than 15% (error of the measurement at 2-sigma). The minimum detection activity for  $^{137}\text{Cs}$  was  $0.1 \text{ Bq kg}^{-1}$ . In order to make sure that the gamma spectroscopy system is working correctly and yields reliable results, the reference material IAEA-375 was measured every second week.

Soil cores from each sampling site were roughly described during the sample preparation and 17 of them (for 46 sub-samples in total) were analyzed for the determination of the main physical and chemical properties. The measured physical parameters were: skeleton content (%), bulk density ( $\text{kg m}^{-3}$ ), particle size distribution (%) measured by the wet sieving method for sand fractions and by the sedimentation method (using the SediGraph 5100 Particle size Analysis System) for the silt and clay fractions. The chemical parameters analyzed were:  $\text{pH}_{\text{H}_2\text{O}}$ , total organic (TOC) carbon content (%) measured by the RC612 Multiphase Carbon and Hydrogen/Moisture Analyzer, total nitrogen content (TN %) measured by the LECOCHN-1000 Carbon, Hydrogen and Nitrogen Analyser.

### 2.2.5 Conversion of $^{137}\text{Cs}$ measurements in estimates of soil redistribution rates

The profile distribution model is the most used for undisturbed stable soils, where the depth distribution of  $^{137}\text{Cs}$  shows an exponential decline with depth that may be described by the following function (Walling and Quine, 1990; Zhang et al., 1990):

$$A'(x) = A_{\text{ref}}(1 - e^{x/h_o}) \quad (1)$$

where:

$A'(x)$  = amount of  $^{137}\text{Cs}$  above the depth  $x$  ( $\text{Bq m}^{-2}$ )

$x$  = depth from soil surface expressed as mass between top and actual depth ( $\text{kg m}^{-2}$ )

$A_{\text{ref}}$  =  $^{137}\text{Cs}$  reference inventory ( $\text{Bq m}^{-2}$ )

$h_o$  = profile shape factor ( $\text{kg m}^{-2}$ ), it is a coefficient describing the rate of exponential decrease in inventory with depth, for a soil profile in uncultivated site.

If it is assumed that the total  $^{137}\text{Cs}$  fallout occurred in 1986 and that the depth distribution of the  $^{137}\text{Cs}$  in the soil profile is independent of time, the erosion rate  $Y$  for an eroding point (with total  $^{137}\text{Cs}$  inventory  $A_u$  ( $\text{Bq m}^{-2}$ ) less than the local reference inventory  $A_{\text{ref}}$  ( $\text{Bq m}^{-2}$ )) can be expressed as (Walling and Quine, 1990; Zhang et al., 1990):

$$Y = 10/(t-1986) \times \ln(1 - X/100) \times h_o \quad (2)$$

$Y$  = erosion rate ( $\text{Mg ha}^{-1} \text{ yr}^{-1}$ )



$t$  = year of sampling

1986 = because in Valle d'Aosta Region the contribution of Chernobyl wet deposition was the major part of the global inventory (84%)

$X$  = % reduction of  $^{137}\text{Cs}$  total inventory in respect to the local  $^{137}\text{Cs}$  reference value (defined as:  $(A_{\text{ref}} - A_u)/A_{\text{ref}} \times 100$ )

The profile distribution model is simple and easy to use. However, this model involves a number of simplifying assumptions and does not take into account the time dependent nature of the  $^{137}\text{Cs}$  fallout migration in the soil and the progressive evolution of the depth distribution of the  $^{137}\text{Cs}$  within the soil profile after deposition from the atmosphere. Consequently, it is likely to overestimate rates of soil loss if long time periods are investigated (Walling and He, 1999). However, in our study area with  $^{137}\text{Cs}$  input predominantly (>80%) from Chernobyl (1986), the application of this model is most suitable because we can assume that the time span for migration of  $^{137}\text{Cs}$  is too short. Statistical analysis of the data was carried out using SPSS 17.0 (SPSS Inc., Chicago, 15 IL).

### 3 Results and discussion

#### 3.1 Assessment of snow related soil deposition rates using the conventional approach

In the two considered events (years 2009 and 2010) volume and areas of the avalanche deposits were comparable. The area was 10 times smaller than the one registered as the maximum event in the Regional Avalanche Cadastre (6.5 ha) (Fig. 1). The measured mean depths of the two events were similar ( $1.36 (\pm 0.20)$  m and  $1.68 (\pm 0.22)$  m) and we used the average value of  $1.52 (\pm 0.21)$  m, rounded down to 1.50 m for the calculation of the inner body volume. In the 2010 event the amount of sediment transported was seven times higher than in the 2009 event. In both years most part of the sediments was concentrated in the upper centimetres of the snow deposit, as shown also in other studies (Jomelli and Bertran, 2001; Heckmann, 2002). In the superficial part of the avalanche deposit the sediment concentration was more than four hundreds times higher than in the inner body with a mean value of the two events equal to  $467.53 \pm (95.31) \text{ kg m}^{-3}$  and  $1.04 (\pm 0.20) \text{ kg m}^{-3}$ , respectively. Also the sediment load from the surface layer ( $60386 (\pm 12305) \text{ kg}$ ) was five times higher than the load derived from the inner body ( $11950 (\pm 2367) \text{ kg}$ ). Sediment concentration and load values were in the range or even higher than the values reported in literature (Ackroyd, 1987; Jomelli and Bertran, 2001; Heckmann, 2002). The snow related soil deposition rates, estimated by the avalanche deposit data, were  $27.5 (\pm 3.1)$  and  $161.0 (\pm 34.8) \text{ Mg ha}^{-1} \text{ event}^{-1}$ , for 2009 and 2010 event, respectively. These values were comparable to data (9.7–240.8

Mg ha<sup>-1</sup> event<sup>-1</sup>) referred to an avalanche path located close to the study area (Freppaz et al., 2010). The soil accretion values, resulting from the sediment accumulation, were 2.29 ( $\pm$  0.26) and 13.41 ( $\pm$  2.90) mm event<sup>-1</sup>, for 2009 and 2010 events, respectively (Table 1). Considering that one avalanche event occurred per season, we can assume that these data are yearly rates of snow related soil deposition and they are in the range of values (0.3-4.8 mm yr<sup>-1</sup>) reported in the previous literature (Caine, 1969a; Luckman, 1971 and 1978).

The pit dug in the avalanche body to sample the snow vertically down to the ground (Fig. 2) confirmed that the sediments was concentrated in the upper few cm of the snow deposit and homogeneously distributed in the inner body ( $0.25 \pm 0.04$  kg m<sup>-3</sup>) revealing a good vertical mixing of sediments.

Regarding the spatial distribution of sediments at the deposit surface, the geostatistical analyses (GIS) didn't show a clear pattern. The total absence of a sediment distribution pattern in the run-out area might be explained by the morphological characteristics of the avalanche path and by the kind of avalanche (full-depth slab, wet snow). The avalanche path is very steep and channelled in the lower track area and suddenly the steepness decreases in the run-out area, so that we could hypothesize that the heavy mass of wet snow suddenly stopped without the space and power necessary to run more and to redistribute the sediments horizontally.

### **3.2 Assessment of snow related soil erosion rates using the conventional approach**

In the considered events (2009 and 2010) the avalanche release and track zones had almost the same shape and the area was equal to about one third of the area of the maximum event reported in the Regional Avalanche Cadastre (18.1 ha) (Fig. 1). The soil erosion rates, estimated assuming that the source of the total sediment load was uniformly distributed over the entire avalanche release and track areas, were 3.7 ( $\pm$  0.4) and 20.8 ( $\pm$  4.5) Mg ha<sup>-1</sup> event<sup>-1</sup>, corresponding to a soil layer of 0.31 ( $\pm$  0.04) and 1.73 ( $\pm$  0.37) mm, respectively for 2009 and 2010 events (Table 1). As for the deposition rates we can consider these data as yearly rates of snow related soil erosion. The erosion rates were comprised into the wide range (1–100 Mg ha<sup>-1</sup> yr<sup>-1</sup>) reported by Bozhinskiy and Losev (1998) and referred to the annual removal of mineral material caused by avalanches in areas of Russia at different elevation zones.

### **3.3 Soil characterization of reference site and evaluation of the <sup>137</sup>Cs baseline level**

In the reference site (RS) the upper horizons showed a well developed polyhedral and granular structure, while the subsoil had a weaker structure and a higher skeleton content (Table 2).

The mass activity in the first 3 cm ranged between 28.8 and 72.0 Bq kg<sup>-1</sup> and exponentially decreased to 8.7–16.1 Bq kg<sup>-1</sup> in the 9–12 cm soil increment. No caesium was detected below the 12 cm layer. The reference site had a mean total <sup>137</sup>Cs mass activity equal to 141.6 Bq kg<sup>-1</sup>. Considering the average areal activity density (kBq m<sup>-2</sup>), calculated on the bases of the soil density, most of <sup>137</sup>Cs (87%) was accumulated in the upper 9 cm (Fig. 3). The total <sup>137</sup>Cs areal activity of the reference inventories ranged from 2.62 to 4.45 kBq m<sup>-2</sup>. The average value of 3.59 kBq m<sup>-2</sup>, with a coefficient of variation (CV) of 22.3 %, corresponded to the caesium baseline level, which was the residual amount left from the historical <sup>137</sup>Cs fallout inputs, in absence of erosion or deposition. This baseline level was within the range (0–6000 Bq m<sup>-2</sup>) of caesium concentrations in the soil, reported for the study area after a measuring campaign conducted by ARPA Valle d'Aosta between 2001 and 2004 (Agnesod et al., 2006). Moreover the coefficient of variation (22.3 %) was within the range of results reported in other studies carried out on reference sites (Sutherland, 1991, 1996; Owens and Walling, 1996).

### **3.4 Assessment of soil deposition rates using <sup>137</sup>Cs approach and comparison with the conventional approach**

The deposition areas (UDA and LDA) and the site above snow bridge (ASB), compared with the reference area, showed a clear sedimentation process; the average deposition rate in ASB (8.9 Mg ha<sup>-1</sup> yr<sup>-1</sup>), even if not significantly ( $p=0.094$ ), was globally lower than in the deposition areas (average of 11.7 Mg ha<sup>-1</sup> yr<sup>-1</sup>) (Fig. 4). In UDA the average total areal activity was 12.76 kBq m<sup>-2</sup> with a maximum activity in the 3–6 and 21–24 cm soil layers (Fig. 5). In LDA the total areal activity was 13.33 kBq m<sup>-2</sup> with a greater activity in the upper soil horizons (0–15 cm), and the maximum in the 10–15 cm soil layer, equivalent to 2.70 kBq m<sup>-2</sup>. In ASB the <sup>137</sup>Cs activity was maximum in the upper soil layer, with an exponential decrease towards depth (Fig. 6) and a total average areal activity of 9.37 kBq m<sup>-2</sup>. The sedimentation rates in the avalanche deposition sites were estimated equal to 12.1 Mg ha<sup>-1</sup> yr<sup>-1</sup> (UDA) and 12.6 Mg ha<sup>-1</sup> yr<sup>-1</sup> (LDA) (Fig. 4). Taking into account the average bulk density of the different soil layers, the soil accretion was estimated at 1.05 mm yr<sup>-1</sup> and 0.93 mm yr<sup>-1</sup>, equivalent to a total deposition during the 24-year period (1986–2010) of 25.10 and 22.40 mm, in UDA and LDA respectively.

The yearly sedimentation rates measured by caesium were lower but comparable with the estimates done using data from the avalanche deposit area (27.5 and 161.0 Mg ha<sup>-1</sup> event<sup>-1</sup> with an estimated soil accretion ranging between 2.29 and 13.41 mm event<sup>-1</sup>). In LDA the effective <sup>137</sup>Cs value could be likely higher as the sampling points were chosen in a rather steep area in the deposition zone (greater than 15°), where it wasn't possible to exclude soil erosion for example by runoff, but not

disturbed by human activity. Moreover, these results, both from conventional and  $^{137}\text{Cs}$  approaches, are higher (about ten times) than the values reported by Lana-Renault et al. (2011), who, in a field experiment in a Mediterranean high mountain catchment, found a yearly sediment yield, due to runoff processes, ranging between 2.1 and 2.9  $\text{Mg ha}^{-1} \text{ yr}^{-1}$  (43-35 % of which due to snowmelt runoff).

In ASB the associated soil accretion rate was  $0.84 \text{ mm yr}^{-1}$ , equivalent to a total deposition during the 24-year period (1986–2010) of 20.23 mm. In the upper (UDA) and lower (LDA) deposition areas the soils, compared to all the other sites, were characterized by the highest skeleton content (52 %) and bulk density ( $1277 \text{ kg m}^{-3}$ ), while in the area above the snow bridges (ASB) soils had lower skeleton content (average of 19 %) and bulk density (average of  $1003 \text{ kg m}^{-3}$ ) (Table 2). Considering the even lower amount of skeleton (12 %) found in the upper soil horizons of ASB, it seems that here mainly fine particles accumulate. In early spring 2011, just after the snow melt and before any rain storms, a clear sedimentation of fine particle was visible (Fig. 7). Therefore in ASB, usually not interested by ground avalanches, we can assume that, besides the runoff processes, also snowmelt and snow gliding, in this case without enough force to transport stones, may actively contribute to the total soil erosion.

### **3.5 Assessment of soil erosion rates using $^{137}\text{Cs}$ approach and comparison with the conventional approach**

In the avalanche release (RA) and track areas (TA) the soils were frequently disturbed with the removal of the upper horizons and the exposure of the subsoil, while in the snow bridges area (SB) no evident erosion was present. Looking at the properties of the upper horizons in the reference site (RS) and in RA, TA and SB, the soil density ranged between  $659 \text{ kg m}^{-3}$  in SB and  $1073 \text{ kg m}^{-3}$  in RA and the skeleton content is the highest in RA (41 %) and the lowest in RS (5 %). The high percentage of skeleton content, except for RS, could confirm that in RA and TA the deeper horizons became exposed to the surface after the erosion of the upper horizons, while in SB the soils were more likely disturbed during the construction of the snow bridge. The organic carbon content is higher in RS (4.1 %) and SB (4.9 %) and lower in RA (3.3 %) and TA (3.8 %), showing that in SB, where also the soil density is the lowest, the vegetation cover seems to have better stabilized the soil (Table 2). The mean  $^{137}\text{Cs}$  activity in RA was  $37.3 \text{ Bq kg}^{-1}$ , while values of  $74.0 \text{ Bq kg}^{-1}$  and  $61.4 \text{ Bq kg}^{-1}$  were found in TA1 and TA2 transects, respectively. In SB the mean  $^{137}\text{Cs}$  activity was equal to  $62.8 \text{ Bq kg}^{-1}$ , with an increasing trend down the slope. Converting the  $^{137}\text{Cs}$  inventories to soil erosion rates, using the profile distribution model, the values confirmed that in these areas erosion phenomena occurred, except for the point at lowest altitude in the SB transect. The SB site

shows a decreasing trend in soil erosion rates from the highest to the lowest sampling point, with  $-30.0 \text{ Mg ha}^{-1} \text{ yr}^{-1}$  and  $+0.3 \text{ Mg ha}^{-1} \text{ yr}^{-1}$ , respectively (Fig. 8). This phenomenon could be due to the protection effect of the avalanche defence structures. It seems that snow bridges have also an influence on soil erosion processes, allowing the colonization by shrubs and larch seedlings. In the other transects the erosion values are more variable and without any trends along the slope. Considering the mean erosion rates in the different areas no significant differences ( $p < 0.05$ ) were found (Fig. 9) even if a lower value was observed in the TA ( $8.8 \text{ Mg ha}^{-1} \text{ yr}^{-1}$  vs.  $13.4 \text{ Mg ha}^{-1} \text{ yr}^{-1}$  in RA and  $12.6 \text{ Mg ha}^{-1} \text{ yr}^{-1}$  in SB). These values are in the range of yearly values ( $7\text{--}37 \text{ Mg ha}^{-1} \text{ yr}^{-1}$ ) determined, also with the caesium-137 method, by Konz et al. (2009, 2010) for steep alpine slopes in the Central Swiss Alps, which are also partly heavily affected by avalanche activity. Taking into account the mean bulk density of the different soil layers, the erosion rates ranged between  $0.97$  and  $1.90 \text{ mm yr}^{-1}$  in TA and SB, equivalent to a total soil erosion, during the 24-year period (1986–2010) of  $23.29 \text{ mm}$  and  $45.71 \text{ mm}$ , respectively. The  $^{137}\text{Cs}$  results are comparable to the erosion values estimated by data from the deposits of the two full-depth slab events considered in this study (Table 1): in fact the average basin mass removal of the two events was  $12.3 \text{ Mg ha}^{-1}$  with an estimated soil removal equal to  $1.02 \text{ mm}$  ( $0.31$  and  $1.73 \text{ mm event}^{-1}$ ). Moreover the estimates of erosion due to the avalanche activity are significantly greater than the erosion rates related only to runoff processes during the growing season. For example Konz et al. (2010) found values in the range of  $0\text{--}68 \text{ kg ha}^{-1} \text{ mo}^{-1}$ , and also other studies confirm low erosion rates during the vegetation period (e.g. Felix and Johannes, 1995; Simonato et al., 2002), even if the effective magnitude of this process depends strongly on the amount of the precipitation during the summer period (Felix and Johannes, 1995).

#### 4 Conclusions

Alpine regions are particularly susceptible to soil erosion because of their extreme climatic and topographic conditions. Avalanche paths are often places where soil is redistributed by snow movements (ground avalanches and snow gliding), as confirmed by the significant amount of sediments found in the deposits of the full-depth slab avalanches considered in this study. The snow related soil accumulation estimated with data from the deposit area is even more intense than the yearly deposition rates assessed with Cs-137. This might be due to the disturbance of the lower deposition area by human activity and/or to the discrepancy between the long-term (since 1986) signal of the Cs-137 method compared to the rates of two single events (2009 and 2010). Also the snow related soil erosion rates estimated from the sediment yield at the avalanche deposit area are comparable to the yearly erosion rates assessed with the Cs-137 method and even if the

comparability between the two techniques is limited by their different time scale, both methods yielded similar magnitudes of soil redistribution rates.

The Cs-137 method also shows that, where ground avalanche doesn't release, the erosion and deposition of soil particles from the upper part of the basin is considerable and likely related to snow gliding, proving the very active contribution of the slow snow movements to the soil erosion. Considering the large natural variability of event sediment loads and the temporal variability of avalanche occurrence, we may expect that when a full-depth slab avalanche occurs, its contribution to the yearly soil erosion rate could be relevant, with values that can be comparable to the total average yearly rates. In fact even hypothesizing that not every year a ground avalanche, with related soil erosion, could have occurred, the important role of avalanche processes in the soil redistribution rates is an undeniable fact. Consequently the soil erosion due to snow movements could be considered the main driving force of soil redistribution in this specific area, with a greater impact in comparison to the runoff processes occurring in the snow-free season. Concluding, the erosion processes due to the "snow season" should be taken into account in the assessment of soil vulnerability on mountain areas as they significantly contribute to the soil redistribution in mountainous areas.

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Table 1 – Main characteristics of the two snow avalanches (2009 and 2010) with the mean concentration values ( $\pm$  SE) of the sediment transported into the Lower Deposition Area (LDA) and with the relative soil accumulation and erosion rates.

AVALANCHE EVENTS		Winter		Winter	
(year)		2008/2009		2009/2010	
Release + track area (m <sup>2</sup> )		50796		60440	
Deposition area (LDA) (m <sup>2</sup> )		6889		7810	
	Deposit	Deposit	Total	Deposit	Deposit
	Surface	Inner Body		Surface	Inner Body
Deposition depth (m)	2.00	1.36 ( $\pm$ 0.20)		2.00	1.68 ( $\pm$ 0.22)
Deposition volume (m <sup>3</sup> )	138	10334	10471	107	11715
Sediment concentration (kg m <sup>-3</sup> )	118.65 ( $\pm$ 12.81)	0.25 ( $\pm$ 0.04)		816.40 ( $\pm$ 177.81)	1.82 ( $\pm$ 0.37)
Sediment load (kg)	16348 ( $\pm$ 1765)	2618 ( $\pm$ 402)	18965 ( $\pm$ 2167)	104425 ( $\pm$ 22845)	21282 ( $\pm$ 4333)
Deposition rate (Mg ha <sup>-1</sup> )	23.7 ( $\pm$ 2.6)	3.8 ( $\pm$ 0.6)	27.5 ( $\pm$ 3.1)	133.7 ( $\pm$ 29.3)	27.3 ( $\pm$ 5.5)
Soil accretion (mm)	1.98 ( $\pm$ 0.21)	0.32 ( $\pm$ 0.05)	2.29 ( $\pm$ 0.26)	11.14 ( $\pm$ 2.44)	2.27 ( $\pm$ 0.46)
Erosion rate (Mg ha <sup>-1</sup> )	3.2 ( $\pm$ 0.3)	0.5 ( $\pm$ 0.1)	3.7 ( $\pm$ 0.4)	17.3 ( $\pm$ 3.8)	3.5 ( $\pm$ 0.7)
Soil erosion (mm)	0.27 ( $\pm$ 0.03)	0.04 ( $\pm$ 0.01)	0.31 ( $\pm$ 0.04)	1.44 ( $\pm$ 0.31)	0.29 ( $\pm$ 0.06)

Table 2 – Main physico-chemical characteristics of the investigated soils (average values).

Site	Total Depth (cm)	Samples	Skeleton (>2mm) %	Sand %	Silt %	Clay %	Bulk Density kg m-3	pH	TOC %	C/N
RS	30-40	(0-18 cm)	5	50.0	40.2	9.8	908	5.4	4.1	10
		(18 cm +)	27	59.9	33.5	6.7	1198	5.3	0.7	7
RA	50-60	(0-9 cm)	41	35.4	38.2	9.7	1073	4.3	3.3	7
		(9 cm +)	47	37.9	34.6	10.8	1073	4.0	1.3	5
TA	50-60	(0-9 cm)	31	32.2	33.2	9.5	908	5.4	3.8	8
		(9 cm +)	42	33.5	30.2	11.4	1137	5.4	1.2	5
SB	40-50	(0-9 cm)	33	39.0	46.9	14.1	659	5.2	4.9	10
		(9 cm +)	38	38.1	42.1	19.8	1085	5.0	2.3	8
UDA	40-50	(0-30 cm)	43	45.5	40.8	13.7	1206	6.0	2.9	9
LDA	40-50	(0-50 cm)	60	56.1	37.3	6.6	1348	8.0	1.6	8
ASB	30-40	(0-18 cm)	12	38.4	47.9	13.7	875	5.0	5.6	9
		(18 cm +)	27	49.0	37.8	13.3	1132	4.8	2.1	6

## Figures Captions

Fig. 1 - Avalanche site called “Torrent de Marais - Mont De La Saxe” - Gray line: perimeter of the maximum event recorded in the Avalanches Cadastre of the Valle d’Aosta Region; white line: release and track perimeters of the events recorded in the seasons 2009 and 2010; gray areas (light and dark): avalanche deposits of the same events (Source: RAVDA-Snow and Avalanche Office); white circles: sample points for Reference Site (RS); black circles: sample points for soil redistribution rates for Release Area (RA), Track Area (TA), Snow Bridge area (SB), Above Snow Bridge area (ASB), Upper Deposition Area (UDA), Lower Deposition Area (LDA).

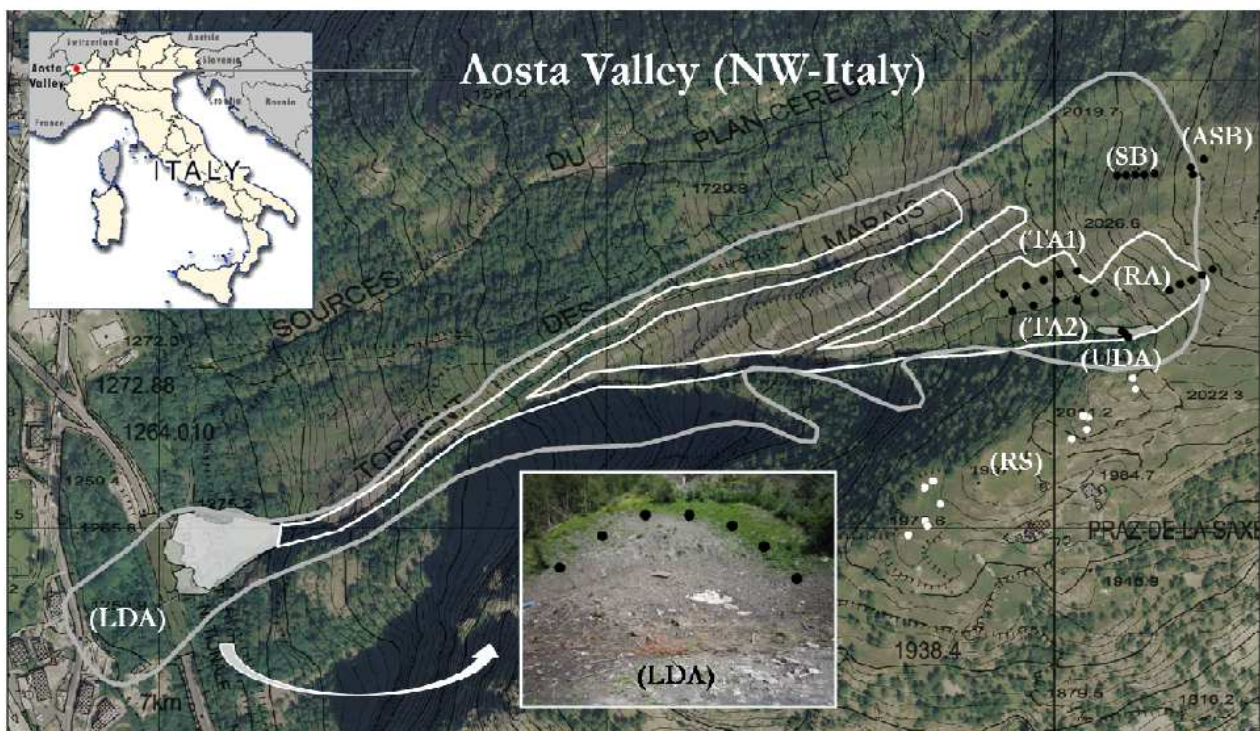


Fig. 2 - Snow deposit in the Lower Deposition Area (LDA): 2008-2009 avalanche event. On the right a particular of the sampling method for the deposit surface (top) and for the sub-superficial snow (bottom); on the left the pit dug into the deposit.



Fig. 3 - Mean depth distribution of  $^{137}\text{Cs}$  ( $\text{Bq m}^{-2}$ ) in the soil of reference site. The sum of the average areal activity density reported for each 3 cm layer is the caesium-137 baseline level.

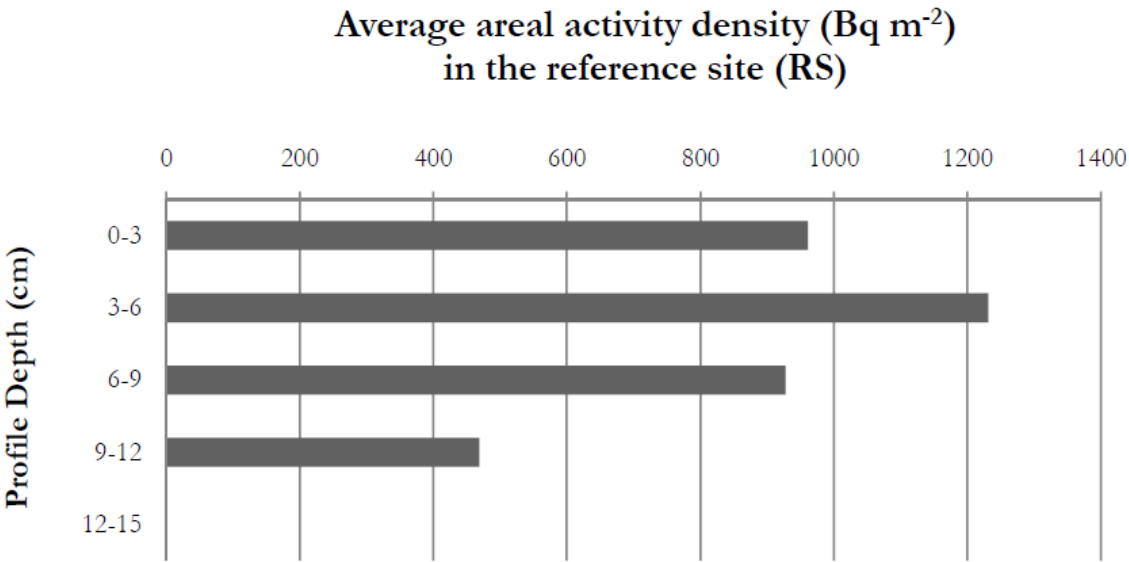




Fig. 4 - Soil deposition rates in the area Above the Snow Bridges (ASB), in the Upper Deposition Area (UDA) and in the Lower Deposition Area (LDA).

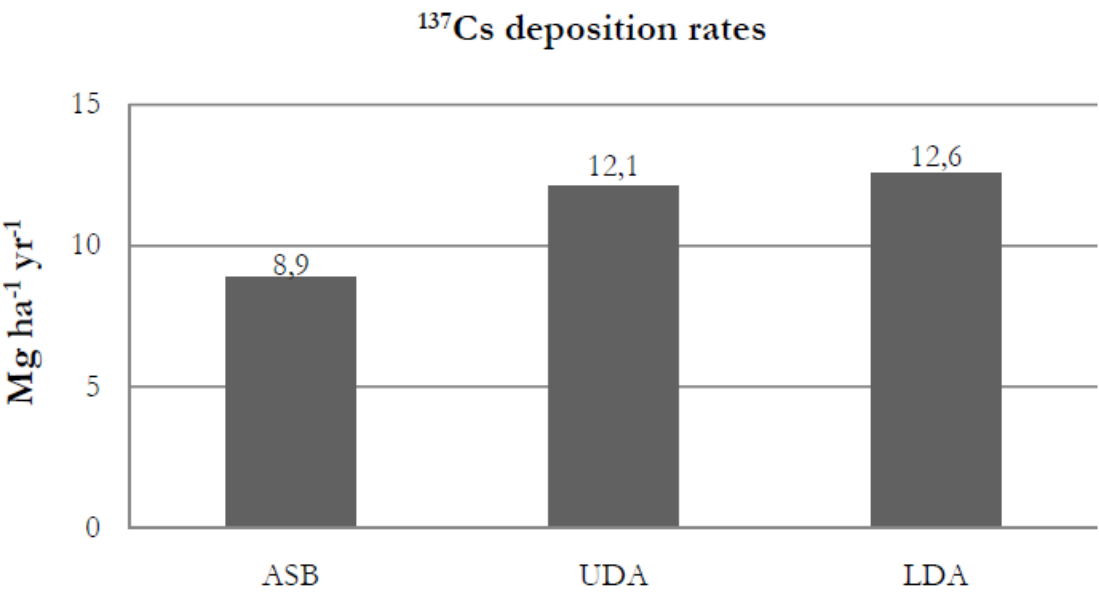




Fig. 5 - Mean depth distribution of  $^{137}\text{Cs}$  ( $\text{Bq m}^{-2}$ ) in the soil of the upper deposition area (UDA).

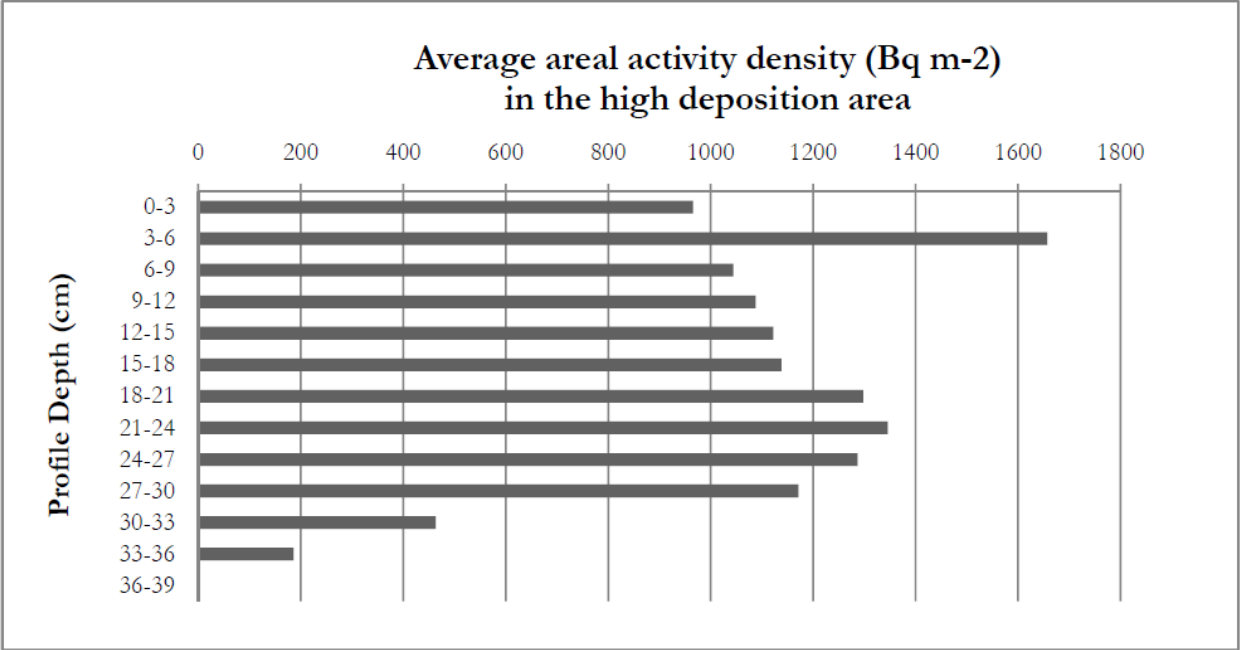


Fig. 6 - Mean depth distribution of  $^{137}\text{Cs}$  ( $\text{Bq kg}^{-1}$ ) in the soil of the area Above Snow Bridge (ASB).

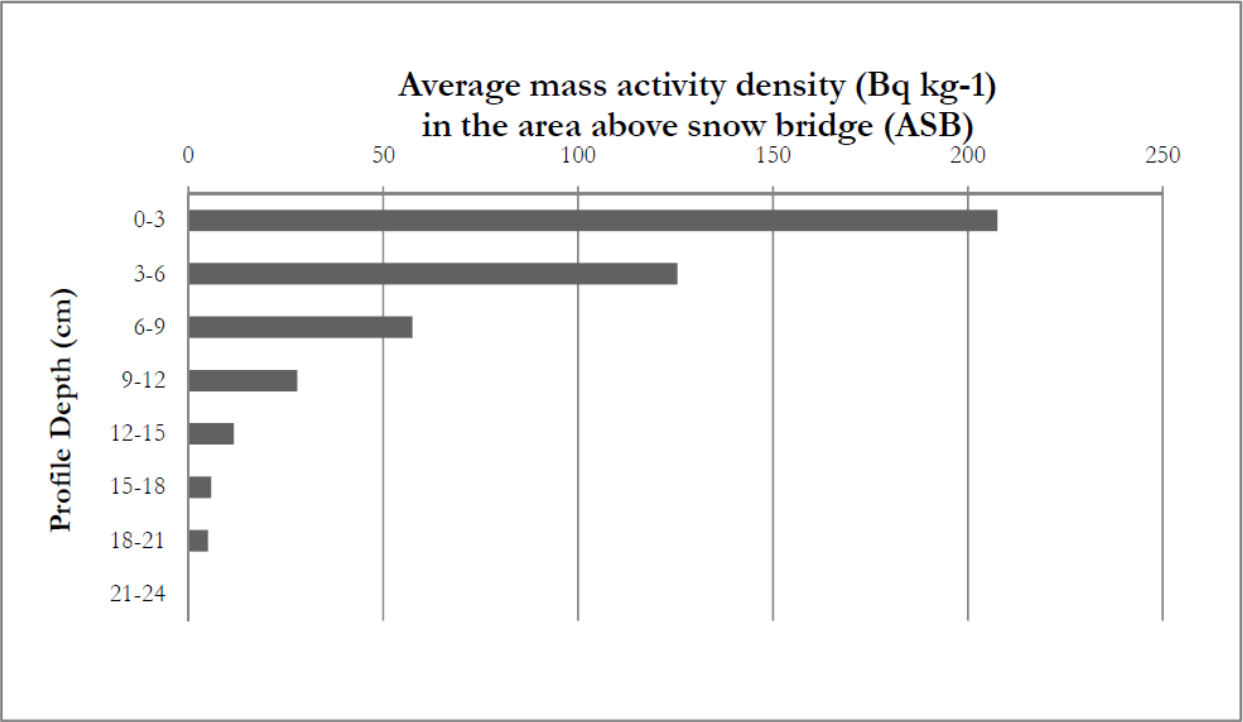


Fig. 7 - Particle sedimentation in the Above Snow Bridges (ASB) area after the snow melting.



Fig. 8 - Soil redistribution rates in all the sampling sites: Snow Bridge area (SB); Above Snow Bridge area (ASB), Release Area (RA); Upper Deposition Area (UDA), Track Area (TA), Lower Deposition Area (LDA). Negative values indicate erosion, while positive ones indicate accumulation. In each transect (SB, RA and TA) the first point in the graph is at highest elevation while the fifth is at lowest elevation.

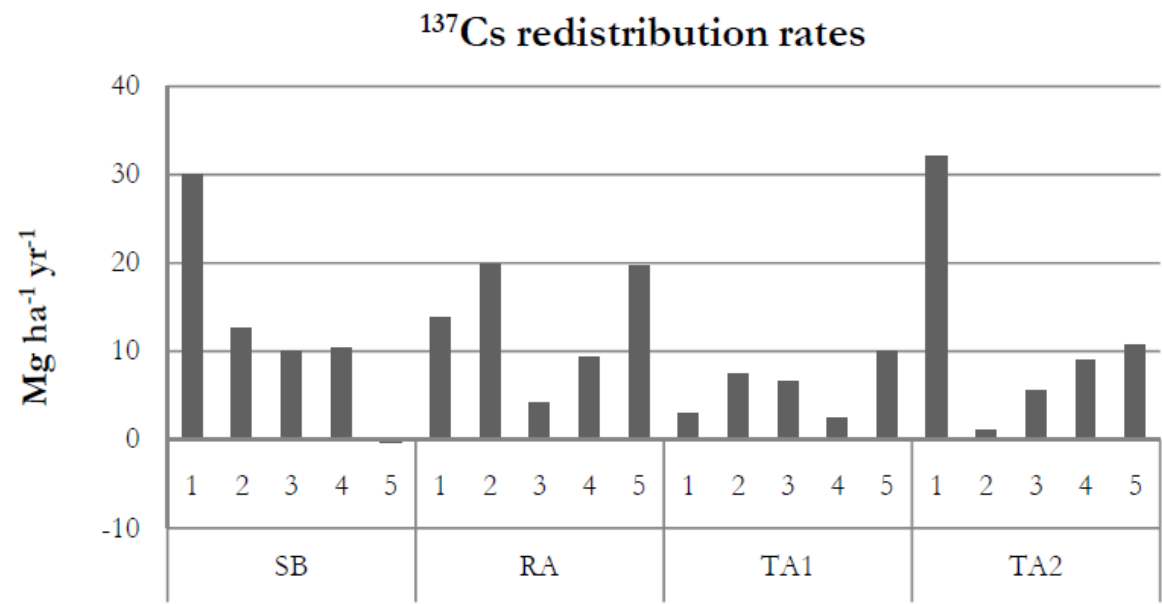


Fig. 9 - Soil erosion rates found in the three sites: Snow Bridge area (SB); Release Area (RA); Track Area (TA), that includes the mean value of all the points belonging from T1 and T2. The dark full circles denote mean values, the blank dark circles represent outliers (i.e. more than 1.5 times the interquartile distance), the black line is the median. Letters a) indicate that there are not significant differences between the means.

